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Two-dimensional modeling of fine sediment transport with mixed sediment and consolidation: Application to the Gironde Estuary, France.

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1 **1. Introduction**

2 Located in the southwest of France, the Gironde Estuary is considered as the largest estuary in Western
3 Europe (180 km in length and 20 km wide, Fig. 1). In the lower and central areas, the Gironde fluvial-
4 estuarine system is characterized by a pronounced turbidity maximum zone (TMZ) that plays an
5 important role in the sediment transport and morphodynamic of the study area (Jalón-Rojas et al., 2015;
6 Sottolichio & Castaing, 1999). In such a highly turbid estuary (1-10 g/L near the bed), the maintenance
7 of the navigation channel is an important issue which arises with the growth of the ship capacity and
8 the predominance of maritime transport. Under this context, Grand Port Maritime of Bordeaux
9 (GPMB) has funded the Gironde XL 3D project, with the support of the European Union (EU). This
10 project aims to develop numerical tools to safely accommodate larger ships by better anticipating
11 channel maintenance requirements and operations. For both objectives, hydrodynamic and sediment
12 transport models are needed over different time scales. On the one hand, the model should be able to
13 predict in real time water levels, flow recirculation and concentrations of suspended particulate matter
14 (SPM) over short periods of time (36 hours) in order to maintain a safe under keel clearance for the
15 ship navigation for given hydrological conditions. On the other hand, the annual time-scale must be
16 considered for channel maintenance operations.

17 Over the years, several numerical models of the Gironde Estuary have been developed. The models can
18 be grouped according to their usage with the first group mainly dedicated to hydrodynamics (Denot et
19 al., 2000; Huybrechts et al., 2012; Laborie et al., 2014). The second group analyzes the turbidity
20 maximum dynamics via three-dimensional models with a fixed bed configuration (Cancino & Neves,
21 1999; Li et al., 1994; Sottolichio et al., 2001, Diaz et al., 2018, van Maanen & Sottolichio, 2018). The
22 last group focuses on two-dimensional, depth-averaged models (2DH) of sediment transport and bed
23 evolution (Chini, 2007; Huybrechts & Villaret, 2013; Van, 2012; Villaret et al., 2011). Huybrechts and

24 Villaret (2013) simulated a 5 year morphodynamic evolution using three sand classes, while Villaret et
25 al. (2011) and Van (2012) did numerical simulations considering cohesive sediment and bed
26 consolidation.

27 The mouth of the estuary is mainly composed of sand (> 90%) while a sand-mud mixture is observed
28 in the intermediate estuary (75 % of mud upstream the P2 station on Fig. 1 according to the GPMB). In
29 the former area, some migrating sand banks also are observed (Kapsimalis et al., 2004). A 2DH
30 modeling approach is selected here as a good compromise between computational cost and hydro-
31 sedimentary processes. Compared to previous works (Huybrechts & Villaret, 2013; Van, 2012), a sand-
32 mud approach is chosen in order to better reproduce the variability of the bed composition. In
33 consequence, even if the model will also be used to predict bed evolution, this paper focuses on the
34 development and the validation of a hydro-sedimentary model. In this way, the model must provide
35 reliable predictions of water levels, current velocities, and SPM concentrations to further set up a
36 Computational Fluid Dynamics (CFD) ship squat model (Ali et al., 2018). To the best of the authors
37 knowledge, few models combining mixed sediment and consolidation processes have been applied for
38 the Gironde Estuary.

39 This paper is organized as follows. In Section 2 the main features of the Gironde Estuary are briefly
40 presented. In Section 3, the hydro-sedimentary mathematical model and its numerical solution are
41 introduced. In Section 4, numerical model calibration and validation are done for different hydrological
42 conditions using measurements of water level, current velocity, salinity, and suspended-sediment
43 concentration collected during the project. In Section 5, a sensitivity analysis for sediment parameters
44 is done to improve the knowledge and the prediction of suspended-sediment concentrations in the
45 Gironde Estuary. Ongoing studies also are discussed in the latter section.

46 **2. Study area**

47 The Gironde Estuary is a macrotidal and convergent estuary with a tidal range varying from 1.5 m
48 during neap tides to 5.5 m during spring tides at the estuary mouth (Fig. 1). The propagation of the tide
49 along the estuary induces an amplification of tidal waves and an asymmetry in the rise and fall of the
50 water level with values of 4 h and 8 h 25 min, respectively. The limit of the tide propagation is located
51 about 170 km upstream of the estuary mouth. Dordogne River and Garonne River contributions to the
52 freshwater discharge are estimated to 35% and 65%, respectively (Sottolichio, 1999). For 2018, the
53 total daily river discharge varied from 133 to 5560m³/s-1 during low and high river discharges periods,
54 respectively. The annual volume of freshwater delivered to the sea is approximately 2.5x10¹⁰m³. The
55 highest velocities are observed in the navigation channel for the intermediate estuary and on the side
56 channel for the lower estuary. Near the bottom, average current velocities are approximately equal to
57 0.75 m/s-1 during spring tides and never exceed 0.5m/s-1 during neap tides (Castaing, 1981).

58 The Gironde Estuary is considered the largest estuary in Western Europe, with the width varying from
59 20 km at the estuary mouth to 3 km for the narrowest sections and with length of 70 km from the Bay
60 of Biscay to the confluence of the Dordogne and Garonne rivers. Based on the bed composition, the
61 estuary can be decomposed into 3 zones comprising (i) a sandy facies in the estuary mouth; (ii) a mixed
62 facies dominated by mud (primarily composed of clays) along the central part, and (iii) a fluvial
63 estuary, in the most upstream parts, characterized by the presence of sand, pebbles, and gravel
64 (Mélières & Martin, 1969; Jalón-Rojas et al., 2015).

65 Fine suspended-sediment observed in the Gironde Estuary are mainly characterized (> 90%) by
66 sediment diameters smaller than 16 µm (Jouanneau & Latouche, 1981). A large part of these fine
67 sediment composes a pronounced Turbidity Maximum Zone (TMZ) with concentrations ranging
68 between 1 and 10 g/LI-1 (Sottolichio & Castaing, 1999). The TMZ formation is mainly due to the tidal

69 asymmetry and partly to the vertical density gradient which maintains suspended sediment in the TMZ
70 (Allen, 1972; Castaing, 1981; Sottolichio et al., 2001). Its location along the estuary migrates between
71 Portets and off the estuary mouth depending on hydrological conditions (Castaing, 1981; Jalón-Rojas et
72 al., 2015).

73 **3. Materials and methods**

74 *1.1. Mathematical models*

75 Hydrodynamics processes are simulated using the continuity and momentum equations solve as the
76 shallow water equations. This model includes parameterizations for the diffusion term and bottom
77 friction. For further details, see Santoro et al. (2017). Bed composition is considered as a mixture of
78 two sediment classes (sand and mud), characterized by their fraction and properties, such as the grain
79 size, the settling velocity, and the concentration. The depth-averaged sediment concentration equation
80 for suspended sediment transport is computed as follows:

$$81 \frac{\partial C}{\partial t} + U \frac{\partial C}{\partial x} + V \frac{\partial C}{\partial y} = \frac{1}{h} \left[\frac{\partial}{\partial x} \left(h \varepsilon_x \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left(h \varepsilon_y \frac{\partial C}{\partial y} \right) \right] + \frac{E - D}{h} \quad (1)$$

82 where C is the depth-averaged concentration of the size class of sediment, in kg/m^3 , U and V are the
83 depth-averaged velocity component along the x and y directions, respectively, in m/s , h is the water
84 depth in m , ε_x and ε_y are the diffusion coefficients along the x and y directions, respectively, in m^2/s . In
85 Eq. 1, the net sediment flux equals the summation of sediment erosion flux E , kg/s/m^2 , and sediment
86 deposition flux D , kg/s/m^2 , computed according to the Parthenadies (1965) and the Krone (1962)
87 formulations, respectively.

88 The bed is discretized with a fixed number of layers in which each layer thickness and sand and mud
89 classes are initialized at time $t = 0$. The state of consolidation is reckoned through a linear relationship
90 between the critical shear stress for erosion of mud ($0.5 < \tau_{ce} < 1.5 \text{ N/m}^2$) and the mud concentration of

91 each layer ($75 < C_{mud} < 500$ g/L). Erosion constants, E_0 , are set constant for all layers. The critical shear
 92 stress for erosion (τ_{ce}) and the erosion rate are computed according to the mud fraction (f_m) (Waeles et
 93 al., 2005). To consider the influence of mixed sediments on bed properties, the authors compute erosion
 94 rates for mud (E_m) and sand (E_s) separately and depending to the regime type (cohesive, non-cohesive,
 95 or mixed). For non-cohesive ($f_m < 30\%$) and cohesive ($f_m > 50\%$) regimes, different formulations for
 96 the critical shear stress for erosion are used accounting for the sediment type (Eq. 2).

$$97 \quad E = f \cdot E_0 \cdot T^a \quad (2)$$

98 Where depending on the sediment type (mud or sand), E_s and E_m are the erosion rates of sand and mud,
 99 $\text{kg/m}^2/\text{s}$, f is the proportion of sand or mud, between 0 and 1, E_{0s} and E_{0m} are the erosion constants of
 100 sand or mud, respectively, $\text{kg/m}^2/\text{s}$, $T = (\tau_b - \tau_{ce})/\tau_{ce}$ with τ_b the bottom shear stress, in kg/m/s^2 , τ_{ce} the
 101 critical shear stress for erosion, and a is a constant equal to 0.5. For the mixed regime ($30\% < f_m <$
 102 50%), a weighted average is used to compute the erosion rates of sand (Eq. 3) and mud (Eq. 4):

$$103 \quad E_s = (1 - f_m) \left(E_{0s} + \frac{E_{0m} - E_{0s}}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit}) \right) T^{a + \frac{1-a}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit})} \quad (3)$$

$$104 \quad E_m = f_m \left(E_{0s} + \frac{E_{0m} - E_{0s}}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit}) \right) T^{a + \frac{1-a}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit})} \quad (4)$$

105 Deposition fluxes of sand and mud are computed by including the deposition probabilities for each
 106 fraction in the Krone equation (Eq. 5). For both classes, the settling velocity W_s is set to a constant
 107 equal to $7.5 \cdot 10^{-4}$ m/s. Following Le Hir et al. (2001) and van Maanen and Sottolichio (2018), a high
 108 critical shear velocity for deposition of mud is applied ($u_{cd}^* = 10$ m/s). Sand and mud suspensions can,
 109 thus, occur during the whole tidal cycle.

$$110 \quad D = \begin{cases} W_s C [(1 - \frac{u^*}{u_{cd}^*})^2] & \text{if } u^* < u_{cd}^* \\ D = 0 & \text{if } u^* > u_{cd}^* \end{cases} \quad (5)$$

111 Where u^* is the shear velocity, m/s. Consolidation processes are accounted for with an iso-
112 concentration multi-layer consolidation model for mixed sediment (Villaret & Walther, 2008; Villaret et
113 al., 2011) that assumes that the vertical flux of sediment between layers is proportional to the sediment
114 mass in each layer, via the use of mass transfer coefficients. This approach was chosen to consider the
115 sediment as a mixture in bed consolidation and differs from previous studies which include bed
116 consolidation using the Gibson theory (Santoro *et al.*, 2017; Thiebot *et al.*, 2011; Van, 2012) with mud
117 only. The bed is discretized with 20 layers characterized by a sediment concentration, a thickness, and a
118 mass transfer coefficient set empirically to reproduce consolidation. During the simulation, only layer
119 thicknesses vary and the mass balance computed for each layer is computed using Eq. 6.

$$120 \frac{M_i(t + \Delta t) - M(t)}{\Delta t} = F_i(t) - F_{i+1}(t) = -a_i M_i(t) \quad (6)$$

121 where i is the layer number, M_i is the mass per unit surface, kg/m^2 , F_i is the sediment flux at the lower
122 boundary of the i -th layer, kg/s.m , and a_i is the mass transfer coefficient s^{-1} .

123 1.2. Model setup

124 Numerical computations are done using the open source Telemac-Mascaret system
125 (www.opentelemac.org). Hydrodynamics are computed using the module TELEMAC-2D which solves
126 the shallow water equations using the finite element method (Hervouet, 2007). This module is coupled
127 at each time step with the module SISYPHE which solves the sediment transport and the bed evolution
128 for both suspended and bedload transport processes. In the current study, only suspended sediment
129 transport processes are considered. The domain is discretized with an unstructured triangular mesh of
130 27,000 nodes that spreads over an area of 2,200 km^2 . The numerical domain extends over the entire
131 estuary from the Bay of Biscay to the limit of the tide influence 170 km upstream from the estuary
132 mouth. The study focuses on the intermediate estuary at the confluence of the Dordogne and Garonne
133 rivers (Fig. 1). Inside this study area, cell lengths vary from 180 m and up to 300 m. Bathymetric data

134 from 2009-2016 collected by the national hydrographic service (SHOM) and the GPMB has been
135 interpolated on the mesh grid. The computational time requires 2 h to simulate 50 d with a time step of
136 10 s (using a work station with 12 cores of 2.4 GHz and 48 Go Ram).

137 Offshore boundary conditions are imposed from North East Atlantic (NEA) tidal atlases (Pairaud et al.,
138 2008) including 46 harmonic constituents for better predictions of water levels (Huybrechts et al.,
139 2012) and upstream boundary conditions are imposed from daily river discharges for both main
140 tributaries. Mean sea level is obtained from the annual average of water levels collected at the Verdon
141 station (Fig. 1) and defined at a value of 0.442 m. This latter value is then applied to the maritime
142 boundary with also a constant salinity of 35 g/L and null values for the SPM. At upstream boundaries,
143 fresh water is imposed while SPM values are taken from Jalón-Rojas et al. (2015).

144 The bottom friction is parameterized over 10 zones with a Strickler formulation, with roughness
145 coefficients varying from 40 to 100 $m^{1/3}/s^{-1}$, obtained from the calibration of water levels and currents
146 (Huybrechts et al., 2012 and Ross et al., 2017). The lower value corresponds to sandy areas of the
147 estuary mouth and is relatively close to the value predicted using the van Rijn formulation considering
148 the bedform influence (van Rijn, 2007; Huybrechts et al., 2012). Higher values of the bed friction
149 imposed in the intermediate and upper estuaries correspond to flat muddy beds.

150 One of the main issues to deal with for mixed sediment (sand-mud or graded sand) is to impose the
151 initial distribution of the mixture over the computational domain and for the bed structure. Due to lack
152 of measurements, the initial distribution is numerically determined assuming a sandy bed (mean
153 diameter = 0.4 mm) at the estuary mouth as suggested by Van (2012). At the intermediate estuary, the
154 mud fraction is predominant in the bed composition with the presence of sand in some locations
155 (Villaret et al. 2010). The bathymetry is defined by the port around 1200 kg/m^3 (McAnally et al., 2016,
156 Xu & Yuan, 2007), corresponding to a concentration layer of $\sim 330 \text{ g/L}$. The fluid mud is initially

157 provided on the navigation channel as proposed by Sottolichio (1999). It is distributed along the first
158 five layers (from 75 to 147 g/L) over an area of 40 km² and has a thickness of the order of 0.5 m. A 2.5
159 m mud thickness is also distributed along layers from 237 to 500 g/L which corresponds to a more
160 consolidated substrate below the bathymetry definition. Simulations are run without considering bed
161 evolution, but the layers update in order to distribute the fluid mud along the estuary.

162 In order to simulate mud consolidation, a calibration procedure was applied to obtain the most
163 representative temporal evolution of the sediment height and associated massive sediment
164 concentrations (Fig. 2). Vertical profiles computed with the consolidation model are adjusted according
165 to the values obtained from settling column experiments (Van, 2012). The maximum number of bed
166 layers has been increased up to 20 layers in order to obtain a better discretization of the bed. Fair
167 agreement between measured and computed values is obtained after the calibration of mass transfer
168 coefficients for each layer. The values of the mass transfer coefficients and sediment concentrations
169 vary from 3×10^{-3} to 1×10^{-6} s⁻¹ and from 75 to 507 g/L, respectively, over layers 1 to 20. These
170 concentrations have been defined according to the experiments of Van (2012) and are in agreement
171 with field observations in the Gironde Estuary (Abril et al., 2000) and other estuaries (McAnally et al.,
172 2007a, b).

173 **2. Numerical model calibration and validation**

174 Numerical model calibration and validation are done by comparing measurements acquired during the
175 project and simulated values for the following parameters : (i) water level ; (ii) current velocity; (iii)
176 salinity; and (iv) suspended-sediment concentration.

177 *2.1. Water levels*

178 Huybrechts et al. (2012) calibrated the bed friction coefficient of the model in order to make the water
179 level difference between measured and simulated lower than 10 cm at the mouth and the central part of
180 the estuary. The calibration and validation have been done for flow currents and water levels obtained
181 during field investigations in August 2006 and October-November 2009. These two events were
182 characterized by low river discharges and calm weather conditions. The hydrodynamic model only
183 included river and tidal forcings. Ten years later, the morphodynamics of the Gironde Estuary may
184 have changed. More recent bathymetric surveys have been performed and can be now considered in the
185 model. It is, thus, necessary to check the validity of the calibration, then to analyze how the model
186 behaves for different hydrological conditions in the upper estuary where the port terminals are located.
187 The calibration update and validation of the hydrodynamics are realized during low ($< 200 \text{ m}^3/\text{s}$) and
188 high river discharges (between 1000 and 2000 m^3/s) for neap and spring tide conditions.
189 During 15 days of low flow conditions (between 08/26 and 09/06/2018), the total river discharge varied
190 from 161 to 235 m^3/s . Comparison between measured and computed water levels shows the ability
191 of the model to predict the tidal amplitude, as well as, the tidal asymmetry with Root Mean Squared
192 Error (RMSE) values up to 18 cm at the Bordeaux station (Fig. 3). Better results are found at the
193 Verdon and Pauillac stations, in the lower and intermediate parts of the estuary (Fig. 1), with RMSEs
194 below 10 cm. Maximum errors are mainly observed during high and low water conditions at the
195 Bordeaux station in the upper estuary. Regression analysis shows a very good fit for every station with
196 coefficients of determination, r^2 , of 0.99 confirming the robustness of the model to predict water levels.
197 However, these accurate predictions of the water level obtained during low river discharges could be
198 degraded during flood periods. Indeed, seasonal variation of river discharges induces the migration of
199 the turbidity maximum (Sottolichio, 1999) which can modify the estuarine bed texture and may have a
200 significant impact on the bed friction (Jalón-Rojas et al., 2018).

201 Differences observed between measured and simulated water levels can come from two sources: a
202 calibration using constant bed friction coefficients with hydrological conditions and/or variation of
203 mean sea level due to tide-surge interactions. To distinguish the influence of the friction calibration and
204 the impact of mean sea level on numerical results, harmonic analysis (Pawlowicz *et al.*, 2002) is also
205 performed on a longer time series at Verdon, Pauillac, and Bordeaux stations (Fig. 4). The lack of data
206 for the year 2018 resulted in the harmonic analysis being done with water levels measured during 2015
207 over a period of 3 months. As expected, the analysis revealed the predominance of the M2 harmonic
208 constituent (principal lunar semidiurnal) in the composition of the tidal amplitude for the predicted and
209 measured values with values varying around 1.55 m (Fig. 4a). The N2 (larger lunar elliptic
210 semidiurnal) and S2 (principal solar semidiurnal) constituent harmonics complete 90% of the signal
211 with amplitudes around 0.3 m and 0.5 m, respectively. Moreover, the amplification of shallow water
212 overtides M4 and M6 is noticed at upstream stations (Fig. 4b-c). Differences between measured and
213 predicted values of amplitude are small and never exceed 7 cm for all harmonic constituents and for all
214 stations. For semi-diurnal constituents (M2, N2, and S2) at the Verdon station, amplitude errors are
215 below 5 cm with a maximum value for the M2 harmonic (4.2 cm).

216 Phase lags are also well reproduced with computed delays of 4 min and of less than a minute for M2
217 and S2 harmonics, respectively (Fig. 4d-e-f). For M4 and M6 harmonics delays increase up to 16 min.
218 For overall harmonics, phase lags do not show significant variations during the tidal propagation. In
219 summary, the harmonic analysis points out that differences in amplitude of the main tidal components
220 are small. For two-dimensional (2D) modeling, this analysis supports the hypothesis of a seasonal
221 calibration of the model. Trial and error methodology as done by Huybrechts *et al* (2012) is not suitable
222 to do regular updates of the bed friction calibration. Therefore, the application of automatic calibration
223 and optimization procedures is more suitable (Smaoui *et al* 2018).

224 2.2. *Current velocities*

225 As also observed for water levels, the accuracy of current velocity predictions decreases in the
226 upstream direction. According to the Fig. 5a-b, results are good enough at Stations P1 and P4 with r^2 of
227 0.79 and 0.90, respectively. However, the robustness of the model decreases at Station P5 where
228 simulations overestimate the strength of ebb currents, particularly during spring tides. For the latter
229 conditions, the measurements describe a well-marked tidal asymmetry characterized by short flood
230 tides and long ebb tides. However, at Station P5 where the measured asymmetry is the strongest, mean
231 velocities of flood and ebb currents are equivalent and equal to 1 and 0.98 m/s, respectively (Fig. 5c).
232 For the same station, the RMSE reaches the maximum value of 0.14 m/s which is not so high compared
233 to other stations where the minimum value is 0.09 m/s. The best results are obtained at Station P1
234 where ebb currents are accurately predicted. For other stations, ebb currents are difficult to predict with
235 an underestimation of the peak ebb velocity and an overall overestimation of velocities during the fall
236 of the tide (Fig. 5b-c).

237 2.3. *Salinity*

238 Validation of salinity predictions used data collected near the bed during mooring surveys and for
239 different hydrological conditions (see Section 3.3). The simulation started during neap tides
240 characterized by low tidal amplitudes of measured salinity (below ~10 psu) for all stations (Fig. 6c-d-
241 e). Conversely, maximum tidal amplitudes are observed during spring tides between days 9 and 13 with
242 a maximum value of 17 psu. However, for the last half of the simulation, the overall salinity tends to
243 decrease in response to a steep increase of the total river discharge of 821 m³/s in 6 d (Fig. 6b-c-d).
244 Despite these changing conditions, the model correctly predicts tidal and fortnightly variations, but
245 tends to underestimate the amplitude of tidal variations, particularly near the estuary mouth at station
246 P1 (Fig. 6c). These differences ranging between 3.12 and 6.7 psu are mainly due to the fact that

247 measurements are taken near the bed, while numerical values of salinity correspond to depth-averaged
248 quantities.

249 A few kilometers upstream, at Station P2, the amplitude is better reproduced, particularly between days
250 8 and 12 (Fig. 6d). However, for some tidal cycles, measured salinity shows rapid changes not captured
251 by the numerical model. This observation could be explained by the presence of a strong vertical
252 density gradient in this area which cannot be reproduced by 2DH model. In the intermediate estuary, at
253 the Station P4, measured salinity decreases rapidly below 5 psu indicating that the salt intrusion limit is
254 close under this hydrological condition (Fig. 6e). In this area, where salinity values are generally low
255 but could increase considerably near the bed, the 2DH model is less accurate but still provides reliable
256 predictions.

257 For the low river discharges period, the highest flowrates are observed between days 29 and 32 with a
258 maximum value of $235 \text{ m}^3/\text{s}$ (Fig. 7b). In this hydrological condition, density stratification is weak and
259 allows accurate predictions, particularly at Station P4 with $r^2 = 0.84$ (Fig. 7d). At Station P1, the model
260 correctly predicts tidal variations of salinity but slightly overestimates salinity during slack waters with
261 a difference up to 2.7 psu for both high and low water slacks, respectively (Fig. 7c). As observed for
262 the flood period, surrounding areas of the salt intrusion limit, where salinity is low, are difficult to
263 predict. At Station P5, conversely to the flood period, the model overestimates salinity, but
264 measurements seem to be wrong and describe a linear signal since day 30 owing to the malfunctioning
265 of the probe (Fig. 7e).

266 2.4. *Suspended-sediment*

267 In order to compare predicted and measured values, a correction factor estimated using a simplified
268 Rouse profile has been applied. The Rouse profile was set using a linear increase of the vertical
269 diffusivity and a reference altitude z_a of 1 m. During the flood period, only two stations were available

270 due to the loss of a probe and the saturation of a sensor at the other one. Hydrological conditions are
271 characterized by higher river discharge periods between days 26-28 and 39-40 (Fig. 8b). However, no
272 clear relation appears with SPM due to the short duration of the simulation.

273 For both Stations P1 and P2, measured and simulated SPM are strongly related to the fortnightly cycle
274 with maximum values observed during spring tides between days 30 and 34 (> 3 g/L for
275 measurements) and null values during neap tides at the end of the simulation (Fig. 8c-d). The dynamics
276 of SPM is reliably predicted by the model excepted at the tidal scale. At this temporal scale, differences
277 between ebb and flood maximum resuspension are not substantial, particularly at Station P1 (Fig. 8c).
278 Although the model tends to underestimate SPM values for both stations, it provides reliable
279 predictions for high river discharges.

280 During the simulation of low river discharges, flowrates are below $300 \text{ m}^3/\text{s}$ and spring and neap tides
281 occurred during days 26-29 and 32-35, respectively. For these hydrological conditions, the TMZ
282 migrates landward near Bordeaux inducing a decrease of SPM in the lower estuary, as observed at
283 Station P2 (Fig. 9c). At this station, the model also accurately predicts the fortnightly dynamics,
284 particularly during neap tides. As observed with high river discharges, tidal variations are still difficult
285 to predict with no clear delineation between flood and ebb resuspensions. For Stations P4 and P5, SPM
286 levels are clearly underestimated by the model by approximately 29 and 125%, respectively, and these
287 underestimations could be explained by a combination of several factors (Fig. 9d-e). First, the model
288 experiences some difficulties to keep suspended particles in the numerical domain over time due to
289 deposition in shallow areas. Second, the numerical reproduction of the TMZ with a 2DH model does
290 not consider the influence of vertical density gradients which act to maintain high suspended-sediment
291 concentrations above the bed.

292 3. Sensitivity analysis

293 A sensitivity analysis is performed to study the influence of several physical parameters such as the
294 critical shear stress for erosion, the settling velocity, the consolidation on depth-averaged SPM (Fig.
295 10) and fluid mud thickness (Fig. 11). The latter parameter corresponds to the sum of the first five
296 layers (< 150 g/L). Results are extracted from the intermediate estuary near the Pauillac station (Fig. 1).
297 For the initial configuration, constant values are chosen for the settling velocity ($W_s = 0.75$ mm/s), the
298 critical shear velocity for deposition ($u_{cd}^* = 10$ m/s) and the Partheniades constant ($M = 2 \times 10^{-3}$
299 kg/m²/s). For the erosion law, a linear relation between the critical shear stress and the mud layer
300 concentration is considered (Sect. 3.1). This choice of relation is similar to those selected by van
301 Maanen and Sottolichio (2018).

302 The first test compares the influence of erosion parameters: the critical shear stress for erosion and the
303 Parthenadies constant (Fig. 10a). Reducing of 20% the critical shear for erosion yields an increase in
304 SPM of 73% and amplifies tidal variations. Conversely, halving the Parthenadies constant induces an
305 overall decrease of SPM of 53% and expands the availability of sediment for consolidation increasing
306 the thickness of the fluid mud layer (Fig. 11a).

307 Similar trends are observed with bed parameters including one simulation without bed consolidation
308 and one simulation with the presence of sand in the intermediate estuary (Fig. 10c). As expected,
309 without bed consolidation SPM concentrations increase and the fluid mud thickness becomes thicker
310 (Fig. 11d). During the first month of the simulation, the fluid mud layer firstly becomes thinner until
311 the mud deposit is completely resuspended and reaches an equilibrium on day 30 (Fig. 11d). However,
312 for simulations covering a period of time of few months, fluid mud deposits in shallow areas are
313 overestimated and reached unrealistic values (> 5 m). For short-term simulations (up to 36 h),

314 consolidation processes can be neglected. Figure 10c also shows the strong influence of the assumption
315 of the sand content. The addition of sand (10% per layer) in the intermediate estuary increases the
316 critical shear stress for erosion, and, therefore, reduces resuspension. Moreover due to consolidation,
317 the fluid mud is transferred to more consolidated layers which tends to increase the sand influence on
318 the first layers. However, imposing 10% of sand in the intermediate estuary and in the entire mud
319 substrate does not seem realistic. Sand is probably located around some banks or between muddy layers
320 which may be exposed to sediment transport according to the turbidity maximum migration or dredging
321 activities. This result points out the importance of monitoring spatial and temporal variations of the bed
322 material composition in order to provide reliable inputs for the model.

323 The last comparisons check the influence of two deposition parameters: the critical shear velocity for
324 deposition and the settling velocity. Reducing by a factor of two these parameters leads to higher
325 suspended-sediment concentrations up to 4.8 g/L⁻¹ and higher tidal variations (Fig. 10b). However, for
326 the settling velocity no significant differences are noticed in fluid mud deposits since the deposition
327 flux is proportional to the concentration of suspended sediment and the settling velocity (Fig. 11b). For
328 the critical shear velocity for deposition, the fluid mud thickness reaches values up to 7-8 cm whereas it
329 reaches 1-2 cm with the baseline simulation (Fig. 11c). A more energetic exchange, thus, is observed
330 between the mud layer and the water column.

331 This parametric study illustrates the sensitivity of the model to the values of sediment parameters.
332 These values are mostly imposed as constants over the computational domain. For the settling velocity
333 it could be interesting to add spatial and temporal variability considering the influence of the
334 suspended-sediment concentration. Similar improvement can be made for the critical shear stress for
335 deposition according to the transport capacity, as suggested by Bi and Toorman (2015).

336 **4. Conclusions**

337 The purpose of this paper is to develop a mixed sediment transport model for the prediction of the
338 under keel clearance in the Gironde Estuary. The current study allows the hydrodynamics and sediment
339 transport to be set up and validated with good accuracy. The harmonic analysis of the astronomical
340 tides reveals (i) a strong distortion of the tidal wave inducing the growth of overtides constituents and
341 (ii) the non-significant effect of tide-surge interactions for annual-scale predictions. Transport for sand-
342 mud mixtures is considered for erosion, deposition, and consolidation processes with two sediment
343 classes (sand and mud) and as a function of the mud fraction. Overall simulations were done with a
344 fixed bed to validate suspended-sediment dynamics. The model was firstly validated for
345 hydrodynamics by comparison with measured water levels and current velocities with high coefficients
346 of determination. Predictions of salinity values also are reliable but show some deviations when the
347 river discharge increases abruptly. Strong vertical density gradients also can explain observed
348 differences. Suspended-sediment concentrations are correctly predicted during flood conditions, but
349 with a less-marked tidal dynamic. When river discharges are low, the model underestimates SPM levels
350 probably due to the difficulty to maintain the TMZ without vertical density gradients. The sensitivity
351 analysis has revealed a strong influence of (i) the settling velocity (ii) the critical shear stress for
352 deposition, and (iii) the chosen erosion parameter for the computation of suspended-sediment
353 concentrations and mud layer thicknesses in the intermediate estuary. This model will be used in the
354 near future to compute the bed morphological changes and analyze the influence of bed load and
355 suspended load transport and dredging/dumping operations in the navigation channel.

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Fig. 1. Location map of the Gironde Estuary (France) and its main tributaries the Dordogne and the Garonne rivers. Measurement stations and tidal gages are represented along the estuary by black squares and black triangles, respectively. The bathymetric chart is computed with the inverse distance weight method and used data from surveys collected at different times from 2005 to 2018.

Fig. 2. Temporal evolution of the relative sediment height z/h (where z is the elevation above the channel bed) for vertical profiles of sediment mass concentration obtained from settling column experiments conducted by Van (2012) (markers) and the semi-empirical multi-layer model (line and markers).

Fig. 3. Time series of simulated (black line) and measured (grey points) water levels at the (a) Verdon, (b) Pauillac, and (c) Bordeaux stations and their respective cross-validations. η_m and η_s correspond to the measured and the simulated free surface, respectively. the simulation is done during a low river discharge period (from August to September 2018).

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Fig. 6. Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged salinity are shown for stations (c) P1, (d) P2, and (e) P4. The simulation is done during a high river discharge period (from March to April 2018).

Fig. 7. Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged salinity are shown for stations (c) P1, (d) P4, and (e) P5. The simulation is realized during a low river discharge period (from August to September 2018).

Fig. 8. Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged SPM are shown for stations (c) P1, and (d) P2. The simulation is done during a high river discharge period (from March to April 2018).

Fig. 9. Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged SPM are shown for stations (c) P1, (d) P4, and (e) P5. The simulation is realized during a low river discharge period (from August to September 2018).

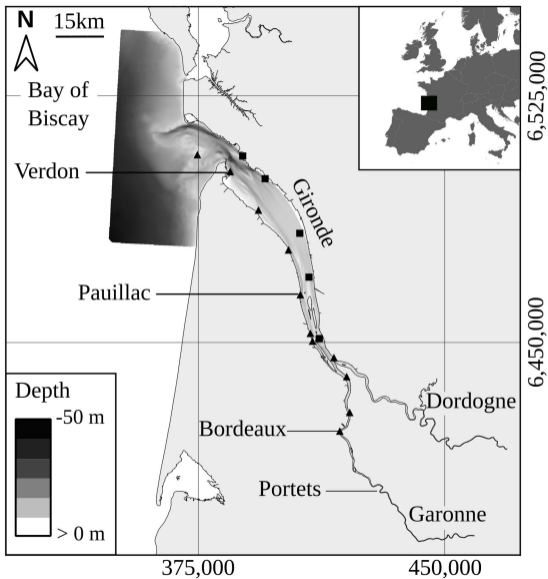
Fig. 10. Time series of depth-averaged suspended particulate matter (SPM) computed by the numerical model with different settings for (a) the critical shear stress for erosion (τ_{ce}) and the Parthenadies

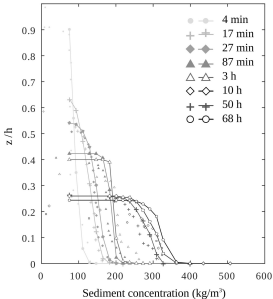
constant (M), (b) the critical shear velocity for deposition (U_{cd}^*) and the settling velocity (W_s) and (c) without consolidation and depending on the sand fraction. Numerical results were extracted near the Pauillac station (Fig. 1) for a period of 60 days.

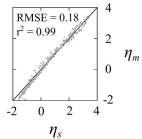
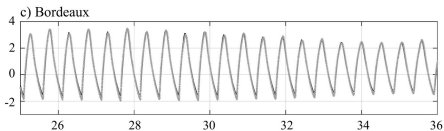
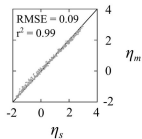
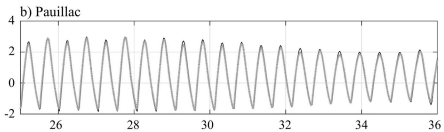
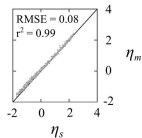
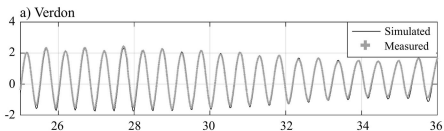
Fig. 11. Time series of the fluid mud thickness computed by the numerical model with different settings for (a) the Parthenadies constant (M), (b) the settling velocity (W_s), (c) the critical shear velocity for deposition (U_{cd}^*), and (d) the consolidation. Numerical results were extracted near the Pauillac station for a period of 60 days.

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Elapsed time since 08/01/2018 (days)

